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Exhumation of (ultra-)high-pressure terranes: concepts and mechanisms

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Abstract. The formation and exhumation of high and ultra-high-pressure, (U)HP, rocks of crustal origin appears to be ubiquitous during Phanerozoic plate subduction and continental collision events. Exhumation of (U)HP material has been shown in some orogens to have occurred only once, during a single short-lived event; in other cases exhumation appears to have occurred multiple discrete times or during a single, long-lived, protracted event. It is becoming increasingly clear that no single exhumation mechanism dominates in any particular tectonic environment, and the mechanism may change in time and space within the same subduction zone. Subduction zone style and internal force balance change in both time and space, responding to changes in width, steepness, composition of subducting material and velocity of subduction. In order for continental crust, which is relatively buoyant compared to the mantle even when metamorphosed to (U)HP assemblages, to be subducted to (U)HP conditions, it must remain attached to a stronger and denser substrate. Buoyancy and external tectonic forces drive exhumation, although the changing spatial and temporal dominance of different driving forces still remains unclear. Exhumation may involve whole-scale detachment of the terrane from the subducting slab followed by exhumation within a subduction channel (perhaps during continued subduction) or a reversal in motion of the entire plate (eduction) following the removal of a lower part of the subducting slab. Weakening mechanisms that may be responsible for the detachment of deeply subducted crust from its stronger, denser substrate include strain weakening, hydration, melting, grain size reduction and the development of foliation. These may act locally to form narrow high-strain shear zones separating stronger, less-strained crust or may act on the bulk of the subducted material, allowing whole-scale flow. Metamorphic reactions, metastability and the composition of the subducted crust all affect buoyancy and overall strength. Future research directions include identifying temporal and spatial changes in exhumation mechanisms within different tectonic environments, and determining the factors that influence those changes.

1 Introduction

Global exposures of high- and ultra-high-pressure (HP and UHP) metamorphic rocks mark regions where crustal materials were subducted to mantle depths and exhumed back to the surface at some point in the geological past (Chopin, 1984; Smith, 1984; Ernst et al., 1997). These rocks provide insight into the long-term dynamics and physical conditions inside subduction and continental collision zones, insight which is not directly measurable or determinable from real-time geophysical data.

Three main groups of crustal materials are subducted, metamorphosed under (U)HP conditions and subsequently returned to the surface: accretionary wedge sediments (producing mainly blueschists), oceanic crust (producing mainly quartz-eclogites in serpentinite-rich subduction channels) and continental crust (producing quartz- and coesite-eclogites during continent–continent collisions (e.g. Guillot et al., 2009). Field observations, geochemical and geochronological analyses, and conceptual, analytical and numerical models suggest that these different tectonic environments exhume high-pressure rocks at different rates and by different, competing mechanisms.

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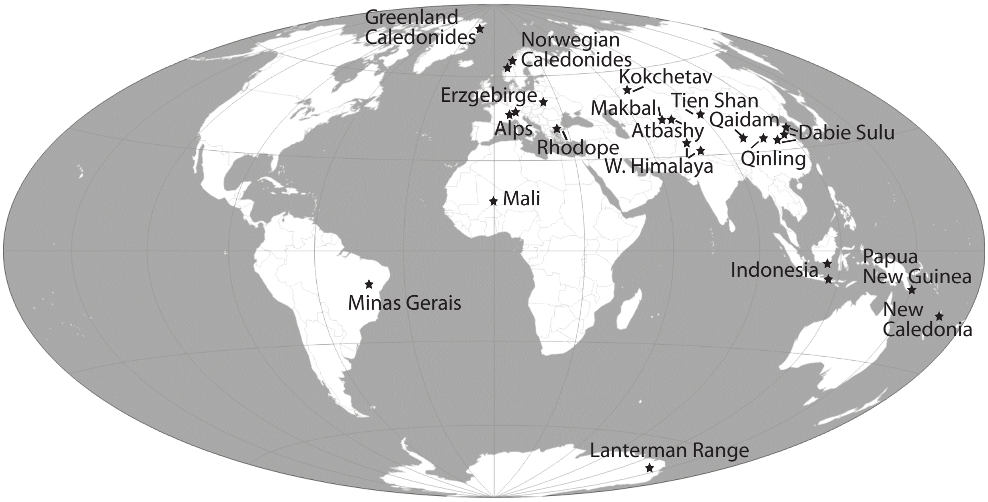

Fig. 1. Map showing the major well-described coesite-bearing, continental crustal, UHP terranes (but not the locations of UHP peridotite bodies). Modified from Chopin (2003), Liou et al. (2004) and Tsujimori et al. (2006).

(U)HP terranes vary in outcrop size, tectonic and structural setting, and age, but they share a number of common features: a commonly upper- to mid-crustal origin, a volumetrically insignificant amount of mafic eclogite with respect to the volume of host felsic gneisses, metasediments or serpentinites, and a post-exhumation structural position close to a plate suture zone (Liou et al., 2004; Fig. 1). The mechanism(s) responsible for their exhumation, and the rate at which they are exhumed, appear to vary significantly between different terranes. Specifically, there appear to be important differences between volumetrically small continental UHP terranes that formed and were exhumed during the early stages of continental collision (e.g. in the Alps and Himalaya) and volumetrically large terranes that formed and were exhumed during later stages of continental collision (e.g. Western Gneiss Region, Norway, and Dabie Sulu, China; Kylander-Clark et al., 2012).

Despite an ever-increasing database of metamorphic, structural, and geochronological data, and increasing insights from high-resolution numerical models, the processes and mechanisms driving the exhumation of continental crustal UHP material, and the cause of the variations seen in different orogens, are still debated. This review firstly summarizes the data recorded by the rocks in many (U)HP terranes around the world. This is followed by a discussion about how subducted crust may weaken and detach from its (possibly still-subducting) substrate and a discussion about the different types of exhumation forcing mechanisms, including those that are *internal* to the exhuming terrane (e.g. buoyancy) and those that are *external* to the exhuming terrane (e.g. tectonic forces). Finally, a discussion about how these different exhumation drivers are accommodated in different tectonic environments, and how the drivers may change within those environments in both space and time, is presented.

2 The record in the rocks

Conceptual models of how continental crust becomes buried to, and exhumed from, depths of up to 120 km are constrained by and based on geological data: metamorphic pressure and temperature determinations, radiometric dating and field structural data.

2.1 Pressures and temperatures

The discovery of natural coesite (the high-pressure polymorph of quartz) in the early 1980s, and diamond (the high-pressure polymorph of graphite) in the 1990s in rocks of unambiguously continental crustal origin such as granitic gneiss, marbles and quartzites, paved the way for the suggestion that buoyant continental crust could be subducted to depths of >100 km and be subsequently returned to the surface (Chopin, 1984; Smith, 1984; Sobolev and Shatsky, 1990). These reports prompted vigorous investigation into the physical processes that could be responsible for continental crustal subduction and exhumation. Terranes containing rocks metamorphosed to UHP conditions have now been documented from most Phanerozoic continental collisional orogens (Fig. 1). The UHP minerals themselves are commonly only preserved as microscopic inclusions armoured in strong hosts such as zircon or garnet, whereas the host rocks generally only tend to preserve evidence for lower pressure conditions. The preservation of these indicative minerals, however, provides evidence for a much deeper subduction history.

Most continental crustal UHP rocks record pressures between 2.5 and 4.0 GPa (Hacker, 2006; Guillot et al., 2009). The most extreme pressures, up to ~ 6 GPa, documented in terranes such as the Kokchetav Massif and Norway's Western Gneiss Region, are commonly recorded in garnet-bearing peridotite bodies hosted within rocks of continental crustal origin, rather than within the host rocks themselves (e.g. Hacker et al., 2003; Zhang et al., 2004; Van Roermund et al., 2002; Vrijmoed et al., 2008, 2009). There is still debate about when and how these peridotite bodies were incorporated into the subducting or exhuming continental crust (e.g. Brueckner et al., 2010, and references therein). Consequently it has been difficult to determine whether the extreme pressures they record correspond to the lithostatic overburden or whether they have experienced significant tectonic overpressures prior to or after their entrainment within the subducting crust (see Sect. 2.1.1).

Most UHP terranes record a period of isothermal or near-isothermal decompression during exhumation until crustal depths (or pressures of ca. 1.0–0.5 GPa) are reached. This period of exhumation has been documented at cm a^{-1} rates (Rubatto and Hermann 2001; Parrish et al., 2006).

2.1.1 Overpressure?

Peak pressures are estimated using petrological reactions involving changes in molar volume. The determined pressures are commonly converted to depth by assuming that the pressure experienced is due only to the weight of the overburden (lithostatic pressure). However, recent field data and numerical modelling results suggest that in some cases, rocks may experience and record tectonic overpressures of up to 1.5 GPa more than lithostatic pressure (Gerya et al., 2008; Vrijmoed et al., 2009; Li et al., 2010).

Orogenic peridotite lenses are common in the metamorphic terranes of most orogens, and record information about how the crust and mantle interact during mountain-building events (e.g. Brueckner et al., 2010). These peridotite bodies commonly record pressures that are much higher than the crustal rocks in which they are found (see Sect. 2.1). In many cases, the timing of attainment of peak pressures in the peridotite bodies is cryptic, and difficult to relate to the timing of peak metamorphism in the host continental crust (e.g. Brueckner et al., 2002; Spengler et al, 2006). In the Svartberget peridotite body, in the northern Western Gneiss Region of Norway, diamond-bearing phlogopite-garnet websterites recording pressures of 5.5 GPa appear to yield Caledonian U-Pb zircon ages of $\sim$ 397 Ma, similar to the timing of leucosome formation in the host felsic gneiss (Vrijmoed et al., 2009). A local melt-induced tectonic overpressure model has been suggested to explain how these two bodies could reach wildly different peak pressures at the same time (Vrijmoed et al., 2009). In their model, the rocks surrounding the peridotite body melted, with the melt confined by stronger, un-melted lithosphere. The pressure increase forced the peridotite to fracture and allowed intruding felsic melt to react with the peridotite to form the fracture-infilling diamond-bearing websterite. Eventually, the pressure was released when the "confining vessel" broke, initiating decompression melting and exhumation.

Numerical modelling suggests that this idea is physically possible, but it remains to be seen whether this is physically plausible in nature. Questions that remain include whether the subducting lithosphere is strong enough to withstand such overpressures, whether melt could cause such high overpressures, and whether the barometers used to provide such high-pressure estimates are accurate.

Numerical geodynamic models of subduction processes suggest that local dynamic tectonic pressures may be greater than lithostatic pressures by up to $\sim$ 0.3 GPa (Gerya et al., 2008; Li et al., 2010). The overpressure estimates in the models are strongly dependent on the choice of model material rheology. Despite advances in the development of self-consistent thermodynamic databases and thermobarometric estimations, pressure determinations in eclogite-facies rocks generally have uncertainties on the order of 0.1–0.2 GPa, similar to the amount of overpressure predicted in some of these models. It is therefore difficult to determine whether high-pressure rocks record pressures greater than lithostatic. This uncertainty in pressures equates to a maximum ca. 10-km uncertainty in burial depth estimates, and it does not greatly affect our understanding of exhumation rates and mechanisms.

More high-precision geochronology and thermobarometric data from other orogenic peridotites in Norway and elsewhere are needed to confirm the Vrijmoed et al. (2009) overpressure hypothesis. Suggestions of tectonic overpressure are currently limited to peridotite bodies, and the question of whether UHP terranes in general have experienced pressures significantly greater than lithostatic remains to be answered.

2.2 Timescales of burial and exhumation

Dating the timing and timescales of the subduction of continental crust relies on the efficient prograde crystallization of the rock-forming and accessory minerals used as geochronometers. Pre-peak chronological information is commonly lost due to (re)crystallization reactions and/or post-crystallization diffusion of the daughter product. Chronometers such as allanite, which crystallize along the prograde path, may record the timing of the near-peak part of the prograde metamorphic path (e.g. Parrish et al., 2006; Smye et al., 2011); and others such as garnet and zircon, which crystallize continuously or episodically over the late prograde–peak–early retrograde part of the metamorphic cycle, may also retain chronological information (e.g. Mattinson et al., 2006; Kylander-Clark et al., 2009).

Data from these chronometers suggest that UHP terranes worldwide spent different amounts of time at UHP conditions: some, like the Kaghan and Tso Morari terranes in the NW Himalaya, spent only a few Ma at (U)HP conditions (Leech et al., 2005; Parrish et al., 2006), whereas others, such as the Dabie Sulu, Quaidam and Western Gneiss Region terranes in China and Norway, may have spent $>$10–20 Ma at UHP conditions (Hacker et al., 2006; Liu et al., 2006; Mattinson et al., 2006; Kylander-Clark et al., 2007; Kylander-Clark et al., 2009).

Geochronological data recording the timescales and rates of exhumation is also variable between different UHP terranes. Depending on the temperature at which the chronometers grew, and the subsequent cooling rate, U-Pb in titanite or rutile, or $^{40}Ar/^{39}Ar$ in muscovite or biotite may be used to provide constraints on the cooling path of the UHP terranes. Rates estimated from the difference between the timing of peak metamorphism and constraints on the cooling path provided by these chronometers vary considerably, from rates on the order of 3–8 $cm\,a^{-1}$ estimated for the Dora Maira and Kaghan terranes (Alps and Himalaya; Rubatto and Hermann, 2001; Parrish et al., 2006), with much slower rates, $< 1\,cm\,a^{-1}$, estimated for the Western Gneiss Region (Kylander-Clark et al., 2008).

The definitive interpretation of thermochronologic data is commonly fraught with difficulty, however, due to uncertainties in determining the timing of mineral crystallization (and hence the proportion of daughter product potentially lost prior to diffusive closure), uncertainties in diffusion parameters, common Pb and excess Ar contamination, and isotopic or crystallographic metastability. These issues are particularly exacerbated in high-pressure terranes due to low permeability, lack of available fluids, high pressures and short metamorphic timescales. Therefore ages recorded by these chronometers should be thought of as providing maximum constraints on cooling ages, as the complications mentioned above tend to increase the yielded apparent age. Exhumation rates estimated from the difference between the peak and exhumation ages may also therefore overestimate the exhumation rate (Warren et al., 2012).

2.3 Structures and microstructures: the record of the exhumation pathway

Different exhumation mechanisms may be determined or inferred by identifying and studying the transport pathways along which exhumation-related motion was accommodated. The exhumation of a UHP terrane as a relatively coherent block or as incoherent flow is generally thought to require the formation of a normal-sense shear zone at the top of the terrane and a thrust-sense shear zone at its base, for at least part of the exhumation history (e.g. Ernst et al., 1997). Alternatively, the exhumation of a UHP terrane by eduction (subduction reversal) geometrically requires a normal-sense shear zone at the upper terrane boundary but a relatively continuous crustal section at the base (Duretz et al., 2012).

Normal-sense shear zones are a common feature of the upper structural boundary of many UHP terranes (e.g. Beaumont et al. (2009) and references therein) but it is generally unclear how much of the deep exhumation these structures accommodated, as their mineralogy and microstructure records mainly crustal-level conditions. The shear zones responsible for transport during the initial stages of exhumation are commonly not preserved or may be cryptically overprinted during exhumation at higher levels (e.g. Young et al., 2011).

Geochronological analysis of fabric-forming minerals in these shear zones may yield an indication of the timescales over which they formed. For example in the Western Alps, Rb–Sr chronology of white micas forming different structural fabrics in the Gressoney Shear Zone, which provides the upper bounds to the HP Piemonte Ophiolite, suggests that the shear zone was operating over a period of ca. 9 Ma (Reddy et al., 1999).

2.4 Short-lived and protracted exhumation events

It has become increasingly clear over the last decade that in orogens where (U)HP material is exposed, exhumation does not always take place during the same orogenic stage in all cases. In the Alps, numerous subduction and collision events have created multiple, diachronously formed and exhumed, (U)HP metamorphic terranes (Schmid et al., 2004; Rosenbaum and Lister, 2005). Recent geochronological and petrological data from one of these terranes, the Sesia Zone, suggest that this terrane was subducted and exhumed twice in the space of a few Ma (Rubatto et al., 2011). In the western Himalaya, geochronological data suggest that different parts of the Indian continental margin were subducted and exhumed at slightly different times during the early stages of the India–Asia collision (Leech et al., 2005; Parrish et al., 2006). In the eastern Himalaya, there is cryptic petrological and geochronological evidence for late-stage (in the last ca. 15 Ma) exhumation of HP rocks (Warren et al., 2011; Grujic et al., 2011).

In both the Alpine and early Himalayan cases, the subduction–exhumation events appear to be discrete, discontinuous and last $<$ ca. 10 Ma, with the terranes residing at (U)HP conditions for < 5 Ma (e.g. Rubatto and Hermann 2001). Exhumation appears to have been mainly driven by buoyancy differences between the subducted crust and the mantle (Liou et al., 2004; Warren et al., 2008b; Beaumont et al., 2009).

In older orogens such as the Norwegian Caledonides or Chinese Dabie Sulu, analytical uncertainty on the geochronological data makes it more difficult to determine whether subduction–exhumation events are continuous and long-lived or overlapping, discrete and short-lived. The data appear to suggest much longer timescales for subduction and exhumation, on the order of ca. 25 Ma (Kylander-Clark et al., 2012 and references therein). These older terranes are generally much larger ($>30\,000\,km^2$ of exposed (U)HP material) and thicker ($\geq$10 km) than in the younger orogens (Root et al., 2005; Hacker et al., 2006). The exhumation of these terranes proceeded tens of Ma after the initiation of continental collision, proceeded more slowly and/or included stalling at lower-mid-crustal depths before final unroofing at the surface (Kylander-Clark et al., 2012 and references therein). The lack of a documented thrust-sense structure underlying the Norwegian Western Gneiss Region (WGR), geometrically necessary for allowing material to return during on-going subduction, and/or in some form of channel flow setting, has led to the suggestion that the WGR at least partially exhumed by "eduction", or reversed subduction (Andersen et al., 1991; Duretz et al., 2012; see Sect. 5.5).

3 Weakening and detaching subducted crust

Examination of exposed continental and oceanic (U)HP terranes shows that they vary along a continuum from being completely pervasively deformed (Terry and Robinson 2004) to consisting of low-strain regions or blocks on a variety of scales separated by high-strain shear zones (Labrousse et al., 2002; Jolivet et al., 2005). Despite their rapid burial and exhumation along a lithosphere-scale shear zone, there are many examples of units within UHP terranes that have accumulated remarkably little strain: well-preserved magmatic fabrics have been reported in (U)HP granites (Wallis et al., 1997), gabbros (Zhang and Liou, 1997; Krabbendam et al., 2000) and pillow basalts (Bearth, 1959; Puga et al., 1995; Oberhänsli et al., 2002). In these localities, all the deformation during burial and exhumation appears to have been accommodated along localized shear zones, many of which formed during exhumation rather than at UHP conditions. The preservation of primary textures also implies low-stress conditions within the subduction zone. Although HP shear zones have been described from many localities (Pognante et al., 1985; Austrheim, 1987; Boundy et al., 1992; Camacho et al., 1997), UHP shear zones are much rarer, although they have been reported from the Dabie Sulu region in China (Zhao et al., 2003). The lack of evidence for structures forming at UHP conditions may be due to metamorphic or deformational overprinting, continuous development and reactivation during exhumation, or an unspecific and/or unrecognizable microstructural record (Stöckhert, 2002; Zhao et al., 2003).

Whilst the location, timing of initiation and mechanism of propagation of the high-strain zones responsible for the detachment of subducted crust and its transport back up to mid-crustal levels of exhumation are poorly constrained, experiments suggest that such shear zones are significantly weaker than the host rock and remain weak once formed (Holyoke and Tullis, 2006). The levels of weakening inferred from natural shear zones associated with UHP rocks may be up to a factor of 100 (Raimbourg et al., 2007). In this section, various possible weakening mechanisms allowing the bulk or localized weakening and hence subsequent detachment of (U)HP terranes are discussed.

3.1 Strain weakening

Experimental strain weakening data are most commonly determined from mono-mineralic starting materials. The data suggest weakening of up to 25–50 % for minerals such as olivine, anhydrite and calcite at low shear strains (Bystricky et al., 2000; Heidelbach et al., 2001; Barnhoorn et al., 2004). Data from strain weakening experiments on quartz (the major mineral in continental crust) is more difficult to interpret as it mainly deforms in the (semi)brittle, rather than ductile, field under laboratory conditions and timescales (Hirth and Tullis, 1992; Hirth and Tullis, 1994; Schmocker et al., 2003).

Outside the laboratory, most rocks are polymineralic and do not show the same rheological behaviour as monomineralic charges in laboratory deformation experiments. In most rocks, several minerals collectively define the rheological properties. Strong minerals may form a load-bearing framework that dominates the strength behaviour, the rheology may be controlled by one or two low-strength minerals that form elongate boudins within the rock, or a single weak mineral may control the rheology while the stronger minerals form clasts (Handy, 1990). Strength vs. composition relationships are highly non-linear (Jordan, 1988; Handy 1990).

Whilst strain arguably causes weakening in natural rocks, the amount of weakening will depend on the mineralogical composition and the amount and scale of heterogeneity of the overall terrane. Neo- and re-crystallization of minerals in shear zones will also change the overall strength of the terrane over time, allowing weak zones to form in different parts of the terrane at different times.

Numerical models of continental collision zones designed to investigate aspects of UHP terrane formation and exhumation commonly apply strain weakening (over and above the non-linear dependence on stress, pressure and temperature) as a numerical proxy for other weakening mechanisms such as grain size reduction, hydration and metamorphic reactions (Babeyko and Sobolev, 2005; Sobolev and Babeyko, 2005; Warren et al., 2008a, c). In these proxies, weakening is applied to the initially homogeneous bulk crust, which weakens by a linear factor when the bulk strain reaches a set threshold. Whilst deformation in the models localizes and partitions into discrete zones during the model runs, the bulk crust also weakens considerably. The extent to which the models mimic the strain partitioning seen on multiple scales in natural rocks is partly dependent on the difference in viscosity between the weakened and un-weakened model crust, and the length-scales over which model and real shear zones develop. Further work is still needed to determine whether the viscosity contrast in the models is reasonable in reality, and also to determine which other factors affect the observed wide variety in strain partitioning scales observed in reality.

3.2 Melt weakening

Partial melting allows rocks to weaken rapidly and effectively. Experiments have shown that there is an exponential decrease in the strength of crustal rocks with increasing proportions of melt (e.g. Rutter and Neumann, 1995; Rosenberg and Handy, 2005). The latter review on the experimental data underpinning the deformational strength of partially melted rocks showed the existence of a non-linear relationship between the amount of melt and overall strength. A significant drop in strength at a melt fraction of ~ 0.07 was noted, corresponding to a transition between a melt-free rock to one containing a "highly interconnected network of melt channels". This significant drop in strength may not be immediately recognizable in the rock record, because the melt films will be thin, and geochronologically useful minerals such as zircon may not grow at this stage. The second, less-pronounced, drop in strength takes place during the solid-to-liquid transition. At this second transition, a "rheologically critical melt percentage" may be able to segregate and pond, forming leucosomes that are recognizable, and minerals that are datable, in the geological record.

Many continental UHP terranes show evidence for migmatization in the mafic eclogite bodies as well as their host rocks, suggesting bulk weakening on a regional scale: see, for example, data from the Western Gneiss Region in Norway (Labrousse et al., 2002, 2011), the Woodlark Basin (Hill et al., 1995; Gordon et al., 2012), the Dabie Sulu region (Wallis et al., 2005), the Kokchetav Massif (Ragozin et al., 2009) and the Greenland Caledonides (Lang and Gilotti, 2007). These studies suggest that conditions suitable for partial melting during the attainment of peak conditions or during exhumation are commonly experienced by UHP terranes. Partial melting must therefore be considered a viable mechanism for the bulk weakening of deeply subducted crust, and a suitable candidate for exhumation initiation.

The importance of partial melting in assisting exhumation and/or driving the initiation of exhumation in continental UHP terranes has been suggested from numerical modelling experiments (e.g. Gerya et al., 2008; Faccenda et al., 2009; Li et al., 2011; Sizova et al., 2012). These studies showed that the bulk weakening, and simultaneous decrease in crustal density associated with partial melting, allow subducted material to detach and exhume. Further numerical simulations of melting in continental collision zones that have not involved such deep subduction of continental crustal material (e.g. Beaumont et al., 2001) have also shown the importance of bulk weakening provided by melting for driving exhumation.

The assessment of the importance of melting to the exhumation of UHP material depends heavily on the timing of melting within the burial–exhumation cycle. Recent high-precision geochronological and geochemical data from the Western Gneiss Region in Norway suggest that melting initiated during high-pressure metamorphism and continued during exhumation-related decompression (Labrousse et al., 2002, 2011). However, data from the D'Entrecasteaux Islands UHP terrane suggests that melting took place after exhumation to lower crustal levels, rather than during UHP metamorphism (Gordon et al., 2012). These data suggest that, at least in this region, melting was a consequence of decompression rather than a driver of exhumation. More work is needed to find evidence, for example, of "nano" or "micro" granites in the UHP record (e.g. Cesare et al., 2009), in order to determine the factors influencing partial melting during the UHP metamorphic cycle and to determine the importance of melting in driving or assisting exhumation. The timing of melting is also critical for the validity of the proposed melt-assisted overpressuring model suggested by Vrijmoed et al. (2009); Sect. 2.1.1.

3.3 Hydration weakening

Hydration of the mantle, leading to its serpentinization, is the most commonly invoked weakening and buoyancy-inducing mechanism for assisting the exhumation of subducted oceanic and continental crust (Guillot et al., 2001; Pilchin, 2005). Oceanic crustal eclogites may reach densities greater than the mantle, so a buoyant matrix is required to assist its exhumation (e.g. Agard et al., 2009). Exhumed oceanic crustal terranes such as the Monviso/Zermatt-Saas in the Alps contain km–cm scale blocks of metabasites within a sheared serpentinite matrix. The serpentinite appears to act as a low-density "carrier" for the higher density oceanic crustal eclogite blocks (Philippot and van Roermund, 1992; Blake et al., 1995; Schwartz et al., 2001).

As well as acting as a buoyancy-increasing agent, serpentinites may also act as a lubricant by forming a mechanically weak zone at the roof of the subduction zone (Guillot et al., 2001). This weak zone assists subduction and subsequent exhumation by decreasing the viscosity of the wall of the subduction channel. Many of the smaller and more geologically recent continental HP and UHP terranes are exhumed structurally beneath the continental suture zone and are spatially associated with mantle-wedge-derived serpentinites (Guillot et al., 2001; Pilchin, 2005; Beaumont et al., 2009). This association appears to be less common in large UHP terranes such as the WGR or Dabie Sulu.

Fluid migration in subducted crust is suggested by the formation of veins bearing eclogite-facies minerals (Philippot and Kienast, 1989; Rubatto and Hermann, 2003), and Caledonian-aged eclogite-facies shear zones bounding Grenvillian-aged granulites in the Bergen Arcs region of Norway (Jolivet et al., 2005). These shear zones suggest that hydration weakening occurred under eclogite-facies conditions. Furthermore, water-filled inclusions associated with plastic deformation structures in UHP omphacite suggest that hydrolytic weakening may play a major role in the weakening of omphacite (Su et al., 2003, 2006). Fluids may additionally assist the formation of relatively weak, foliation-forming metamorphic minerals such as phengite, biotite and lawsonite. The crystallization of these minerals in high-strain zones may help localize deformation and initiate decoupling of buoyant crust from the subducting slab. The formation of discrete weak shear zones may also protect stronger lenses and boudins of UHP material from penetrative deformation, assisting in the preservation of UHP indicator minerals (Guillot et al., 2009).

3.4 Grain size reduction

Minerals in shear zones commonly show a reduction in grain size compared with the original protolith, with an inverse relationship between the amount of strain and the grain size (resulting in the formation of mylonites). The mechanisms responsible for deformation-induced grain size reduction include cataclasis, metamorphic reactions and dynamic recrystallization (De Bresser et al., 2001). Cataclasis is a brittle grain-size-reducing mechanism and is therefore not relevant to the discussion of high-pressure rock exhumation, which initiates in the ductile field.

Eclogite-facies (HP) shear zones have been documented from a number of locations including the Western Gneiss Region and Bergen Arcs in Norway, and the Western Alps (Pognante et al., 1985; Austrheim, 1987; Boundy et al., 1992; Camacho et al., 1997). In the Bergen Arcs, mylonitic eclogite-facies shear zones transect coarser, apparently unaltered, granulite-facies rocks, suggesting that deformation, and arguably fluid, were necessary for triggering the eclogite-forming reactions (Austrheim and Griffin, 1985). What is still unclear is whether deformation triggered the metamorphic transformation or whether fluid triggered the metamorphic reactions, thereby forming a weak zone and hence allowing further deformation and grain size reduction to proceed. Once initiated, deformation was partitioned preferentially into those zones.

Dynamic recrystallization leads to weakening and strain localization by generating a change from dislocation creep (which is insensitive to grain size) to diffusion creep (which is sensitive to grain size; De Bresser et al., 2001). However, significant weakening by grain size reduction may only be possible when caused by syntectonic reaction recrystallization or cataclasis rather than dynamic recrystallization (De Bresser et al., 2001). Alternatively, weakening may only occur if grain growth is inhibited.

Element solubility is enhanced at high pressures. Dissolution–precipitation creep and grain-boundary sliding have therefore also been proposed as likely mechanisms for rock deformation at UHP conditions due to a continuous fluid supply from dehydration reactions and high concentrations of solutes (Stöckhert, 2002). Such mechanisms would imply grain-size-sensitive behaviour and a Newtonian (linear) dependence of strain rate on stress (Rutter, 1983). The expected effective viscosities of rocks undergoing such flow are still poorly constrained by experimental results.

The effect of increasing temperature on decreasing viscosity may be counterbalanced by a temperature-related increase in grain size (Stöckhert, 2002). Further investigation into the causes, and effects on viscosity, of grain size reduction in shear zones is needed, especially in polymineralic rocks.

3.5 Development of foliation

Strain partitioning between the mineral phases in a polymetamorphic rock may lead to the crystallization and/or rotation of planar minerals such as micas. The formation of foliation lowers the bulk strength of the rock by increasing the distance between stronger clasts (Handy, 1990). Foliation development, grain size reduction and/or hydration weakening may all act contemporaneously at different stages of the subduction–exhumation cycle.

3.6 Discussion

This section has summarized ways in which different mechanisms may act to weaken subducted crust, thereby allowing it to detach from the subducting slab and increasing its exhumation potential. These different weakening mechanisms operate at different spatial and temporal scales and at different times during the evolution of the subduction system. They variably depend on pressure, temperature, mineralogy, amount of hydration and any pre-existing heterogeneities within individual lithologies and between different geological units or terranes.

Melt weakening arguably has the greatest and most rapid weakening effect, but many UHP terranes do not appear to have reached suitable PT conditions for melting to have taken place. Hydration weakening and serpentinization are also of great importance in assisting the exhumation of oceanic crustal (U)HP material and probably play an important role in exhuming more shallowly subducted material such as blueschists in accretionary wedge complexes.

More work is still needed to determine which weakening processes dominate at different times and depths within subduction zones, and the amount by which they weaken the crust. It is also important to determine whether any one weakening mechanism is generally responsible for detaching (U)HP terranes from their underlying subducting slab, and/or whether this mechanism may be linked to detachment depth and/or terrane composition.

4 Types of exhumation mechanism

Exhumation of deeply subducted crust may be driven (and/or assisted and/or hindered) by a force *internal* to the exhuming terrane, for example buoyancy (Ernst et al., 1997; Fig. 2a), and/or by a (tectonic) force *external* to the terrane, for example constriction, shear traction or erosion (e.g. Platt, 1993; Figs. 2b–d). It has been suggested that terrane buoyancy is a pre-requisite for exhumation (e.g. Ernst et al., 1997; Warren

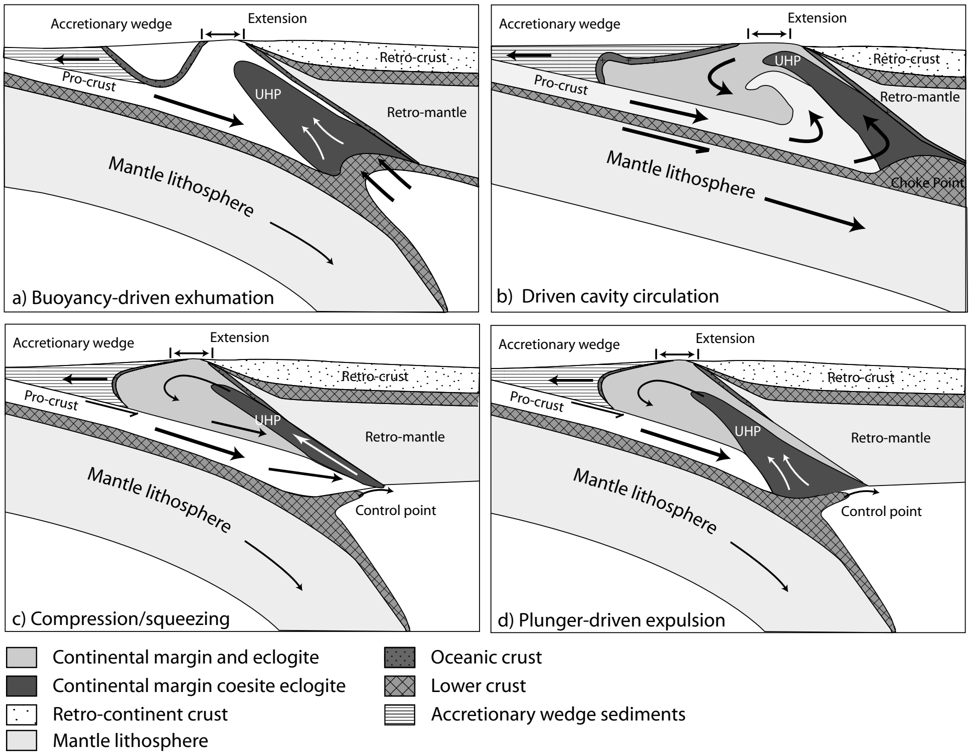

Fig. 2. Schematic diagram showing example styles of exhumation. **(a)** Buoyancy-driven exhumation, where upwards motion is driven primarily by the buoyancy force of the exhuming material; **(b)** Driven cavity flow, where upwards motion is driven by circulatory flow initiated by the traction of the subducting plate; **(c)** Compression/squeezing of weaker material between two stronger blocks (subducting plate and mantle wedge); and **(d)** Plunger-driven expulsion whereby weaker material in the channel is expelled upwards between a choke point at the base of the channel and stronger crust entering the channel. Modified from Warren et al. (2008b).

et al., 2008a, b, c), but it is difficult to assess this claim based on field and geochronological evidence alone. The contribution or dominance of additional assistance and/or resistance from boundary and/or tectonic forces remains unclear.

Scenarios in which external forces can drive exhumation include corner or driven cavity flow environments (Cloos, 1982; Platt, 1993; Warren et al., 2008b; Fig. 2b), extensional collapse (Thompson et al., 1997), compression or squeezing of weak material between two stronger blocks (Thompson et al., 1997; Fig. 2c), plunger exhumation involving the insertion of stronger crust into a channel of weaker material (Warren et al., 2008b, 2011; Fig. 2d) and focused erosion (e.g. Beaumont et al., 2001). Care must be taken to separate the *driving force* from the *tectonic environment* in which that force drives exhumation. For example slab rollback is an environment that changes the force balance within the subduction channel and appears to allow exhumation to proceed (e.g. Lister et al., 2001; Brun and Faccenna, 2008). Scenarios in which exhumation is driven by a combination of forces have also been proposed, e.g. buoyancy in combination with erosion and tectonic processes (Chemenda et al., 1995). These scenarios will be discussed further below.

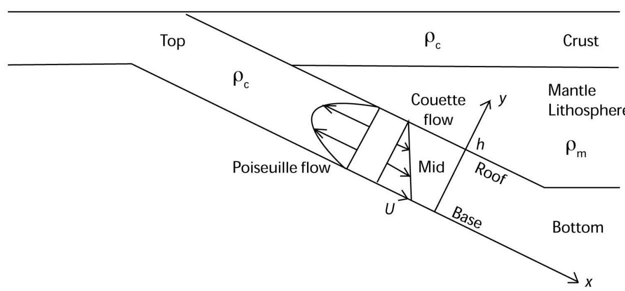

Fig. 3. Schematic diagram showing the nomenclature of subduction/exhumation channel flow behaviour in terms of dominating Couette (subduction) and Poiseuille (exhumation) flows. Figure modified from Warren et al. (2008c).

4.1 The subduction channel

The formation and exhumation (U)HP rocks appears to have most commonly taken place in a subduction channel environment, i.e. a layer of weaker material deforming between a stronger base (subducting slab) and a stronger roof (the base of the upper plate; Fig. 3). In many cases, especially in more recent orogens such as the Alps and Himalaya, exhumation most likely proceeded during on-going subduction. The exhumation potential of buoyant material in a viscous channel which is open at the bottom end and which exhumes against a downwards-directed flow of material at its base has been described in terms of an exhumation number, E, which expresses the balance and competition between the downwards-acting subduction (Couette) flow and the upwards-acting exhumation (Poiseuille) flow:

$$E = (\partial P/\partial x)h^{*2}/\eta_e U, \quad (1)$$

where P is the effective down-channel pressure gradient, x is the horizontal distance along the channel, h^* is a function of channel thickness, η_e is the effective viscosity and U is the subduction velocity (Raimbourg et al., 2007; Warren et al., 2008c; Fig. 3). Although it is simplest to consider a channel of fixed thickness containing material of fixed viscosity, in reality both h^* and η_e are likely to vary within the subduction channel and also over time. Therefore E is also highly, and non-linearly, variable in space and time, and difficult to quantify precisely. The concept of E is useful, however, as it allows the main factors which may contribute to, or hinder, exhumation to be identified.

The depth to which a relatively buoyant continental crustal terrane can be subducted therefore depends on the extent to which the shear traction between it and its surroundings can compete against the buoyancy and/or tectonic forces that act to return the terrane to neutrally-buoyant depths (Ernst et al., 1997). In order to subduct, the subducted crust has to remain strong and/or be subducted rapidly, and/or the subduction channel has to be narrow. In order to exhume during on-going subduction or after subduction has ceased, the subducted continental crust either has to detach from the down-going slab (Raimbourg et al., 2007; Warren et al., 2008a, c), or the downwards-acting traction needs to be removed by, for example, slab breakoff, allowing whole-scale reversal of subduction ("eduction", e.g. Duretz et al., 2012). Exhumation is facilitated by the presence of a wide subduction channel and/or by a slow subduction velocity.

4.2 Internal driver of exhumation: buoyancy

Metamorphic reactions that affect the density and strength of crustal material depend on, amongst others, pressure, temperature, time, deformation and the availability of fluids, including melt. The rate of achievement of stable assemblage equilibrium at any point on the P-T-t path in rocks of differing bulk composition in different tectonic settings is still under debate (e.g. Peterman et al., 2009). Fully transformed metabasaltic eclogite is denser than the mantle at UHP conditions (e.g. Ernst et al., 1997), which may be one of the reasons why remnants of eclogitized oceanic crust are so rarely exhumed. However, continental crust (with an average composition of coesite-bearing granitic gneiss, and $< 10\,\%$ mafic material by volume) is less dense than the surrounding mantle at equivalent conditions (Ernst et al., 1997). Density calculations based on the composition of the Norwegian Western Gneiss Region crust show that even if the entire slab had transformed into dense minerals during subduction (which Krabbendam et al., 2000, and Wain et al., 2001, consider unlikely), it would still have been less dense than the surrounding mantle at UHP conditions (Walsh and Hacker, 2004; Fig. 4).

Whether the felsic continental crust completely transforms into a mineral assemblage stable at eclogite-facies conditions or whether lower pressure minerals persist is an unresolved issue in the study of subducted and exhumed continental crust. Evidence from numerous localities, including the Bergen Arcs and Western Gneiss Region in Norway, and the Western Alps, suggests that granulite-facies rocks can persist metastably through a subduction–exhumation cycle so long as they remain dry and/or undeformed (Wain et al., 2001, and references therein). What is less clear is whether "wetter" continental crust behaves equally metastably under subduction conditions. A recent study of garnets in the quartzofeldspathic gneisses of the Norwegian WGR showed that the cores mostly grew during prograde metamorphism (but only up to ca. 1.7 GPa) and the rims grew mainly during the amphibolite-facies overprint that post-dated UHP metamorphism (Peterman et al., 2009). Very little garnet appears to have grown during the HP-UHP stage, implying that much of the WGR country rock did not transform into eclogite-facies-stable mineralogy during subduction. This observation may also have implications for the geochronology of the HP-UHP stages, especially when using garnet Sm–Nd or Lu–Hf chronology. Zircon, commonly used to date the timing of peak pressure metamorphism, may only have grown during pre-HP prograde or during retrograde amphibolite-facies conditions. Careful trace element analyses of any dated zircon should help to determine the conditions under which it crystallized (Rubatto, 2002; Rubatto and Hermann, 2007).

Similar observations on incomplete transformation of subducted crust have been reported from the Western Alps, where the Brossasco granite in the Dora Maira Massif appears on the macroscale to have remained undeformed and unreactive during subduction and exhumation (Lenze and Stockhert, 2007). Microstructural evidence from quartz suggests that it transformed into coesite and back into quartz during subduction and exhumation. Laboratory experiments suggest that this transition takes place within hours (Lenze et al., 2005). Randomly oriented twins in jadeite and kinbands in kyanite suggest that deformation at UHP conditions was driven by the volumetric change from coesite to quartz rather than any far-field tectonic stress (Lenze and Stockhert, 2007). This evidence suggests that even though some minerals remain unreactive and metastable, the quartz–coesite transition takes place instantaneously, at least on geological timescales.

Sluggish reaction kinetics during the transformation of continental crust in subduction zones will therefore reduce the expected density of the subducted crust. This means that

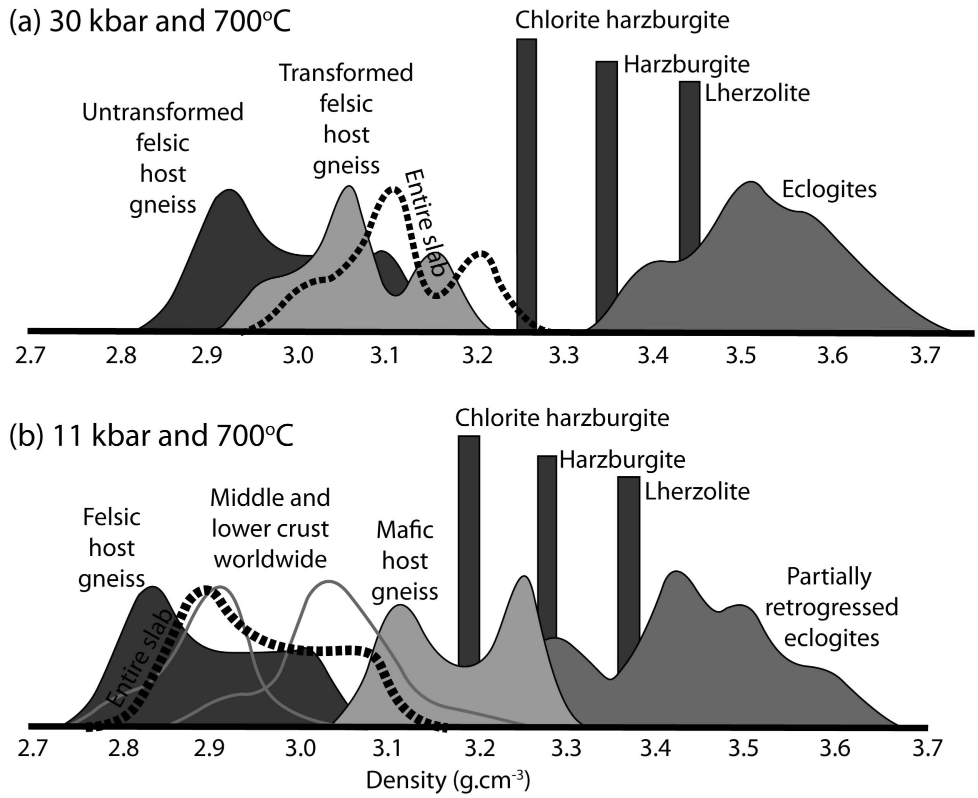

Fig. 4. Figure modified from Walsh and Hacker (2004) showing the calculated densities of the WGR at **(a)** UHP conditions and **(b)** during the amphibolite-facies overprint, using the formalism of Hacker and Abers (2004). Mafic rocks were estimated to form 10 % by volume of the subducted slab.

it will be more difficult to subduct the crust to great depths, unless it is particularly strong and can remain attached to the downgoing slab. As discussed in the previous section, many of the crust-weakening processes require hydration and/or mineral recrystallization, so if the crust is particularly dry, then the decoupling mechanisms may also proceed more slowly. More research is needed into the rate and completion rate of metamorphic reactions and the factors contributing to reaction overstepping and metastability to improve knowledge of material pathways in subduction zones and the relative influence of buoyancy vs. external tectonic forces in driving exhumation.

4.3 External drivers of exhumation

Forces external to the exhuming terrane, such as tectonic forcing or erosion, may be important in driving and/or assisting exhumation. Numerical models of the transition from oceanic subduction to continental collision show that tectonic forcing mechanisms such as plunger expulsion, extrusion and driven cavity (forced return) flow are both physically plausible and may operate at different depths and at different times in the subduction channel (Burov et al., 2001; Warren et al., 2008b). Proving the extent to which these mechanisms drive exhumation in reality, however, requires careful interpretation of structural and geophysical data; metamorphic P-T geochronological data may not provide unambiguous proof.

4.3.1 Plunger expulsion

This mechanism involves the expulsion of weaker (hotter) material due to the insertion of stronger (colder) material into the subduction channel (Warren et al., 2008b; Fig. 2b). Entrance of stronger crust into a subduction channel forces a strong downwards flow of weaker material ahead of it. Exhumation flow is generated when the subducting flow is constricted by a "choke point" at the base of the subduction channel. Numerical models suggest that, in order for the plunger mechanism to operate, the plunger must be considerably stronger than the material in the subduction channel (Warren et al., 2008b). Furthermore, the along-channel pressure gradient driving the return flow must be sufficient to overcome the traction forces at the top of the channel and along the roof of the plunger. For a system subducting material of heterogeneous strength, the plunger model predicts episodic exhumation during the subduction of the stronger units. Based on geophysical evidence for a mid-crustal ramp, this mechanism has been suggested for the exhumation of cryptic lower-crustal granulitized eclogites in the eastern Himalaya (Kellett et al., 2010; Grujic et al., 2011; Warren et al., 2011), but its operation may be difficult to prove in ancient collisional settings.

4.3.2 Driven cavity (forced return) flow

Driven cavity flow is induced by the traction of the subducting material along the base of the channel (Fig. 2d). This traction drives a circulatory or eddy flow that exhumes material out of the channel. Its higher structural level equivalent, corner flow, was proposed in the 1980s to help explain the exhumation of blueschists in subduction zone mélanges (Shreve and Cloos, 1986; Cloos and Shreve, 1988). Using numerical models, Burov et al. (2001) and Warren et al. (2008b) showed that a similar mechanism could also explain the exhumation of more deeply subducted material.

To drive efficient exhumation by driven cavity flow, subducting crust must detach from the subducting plate and accrete in the subduction channel. Exhumation is efficient when the subducting material and accreted channel material are strongly coupled – if not, then the material in the channel will stagnate. Driven cavity flow will exhume material of any density. If the upwards-directed flow is focused into a narrow channel, the exhumation velocity may be faster than the subduction velocity (Gerya and Stoeckhert, 2006). This mechanism can, in theory, produce continuous exhumation over an extended period of time. The operation of this mechanism in fossil subduction zones would be difficult to prove, but it may explain the exhumation of mélanges of blocks recording different P-T-t paths and variable timing of attainment of peak metamorphism.

4.3.3 Erosion

The removal of surface overburden by erosion, especially when focused at the orogenic front, may be an efficient way of assisting and/or driving the exhumation of deeply buried material in combination with tectonic processes (e.g. Beaumont et al., 2001). Numerical models investigating (U)HP exhumation processes suggest that erosion is not required for driving or assisting exhumation from mantle to crustal depths (e.g. Warren et al., 2008b). Rates of exhumation of many (U)HP terranes appear to be on the order of 3–8 cm a^{-1} (Rubatto and Hermann, 2001; Parrish et al., 2006); meanwhile surface processes operate at rates on the order of mm a^{-1}. The difference in rates plus the generally large pressure gaps between the (U)HP and neighbouring terranes suggest that erosion is neither the only nor the dominant driver of exhumation over a large proportion of their history. However, once the exhumed terrane has reached at least mid-crustal levels, a combination of erosion plus low-angle extension faulting may be responsible for exposing the deeply subducted material at the surface.

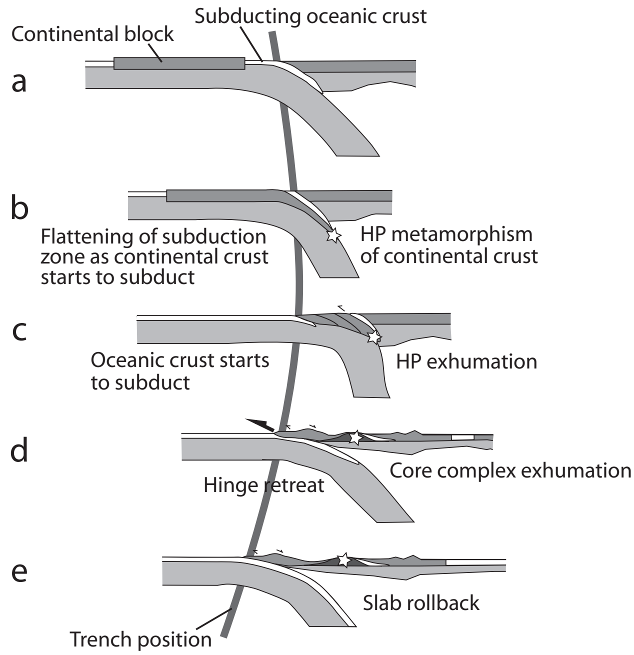

Fig. 5. Figure modified from Brun and Faccenna (2008) showing the interpreted development of slab rollback tectonics and the exhumation of high-pressure rocks in domal culminations within that framework.

5 Tectonic environments facilitating exhumation

5.1 Slab rollback

In order for deeply subducted material to exhume, whether during on-going subduction or after subduction has ceased, the upwards-acting forces must overcome the downwards-acting forces. One way to reduce the traction forces is to reduce the coupling between the upper and lower plates by separation, thereby creating space, and a weak zone, into which the exhuming material can move (Dewey, 1980; Malusà et al., 2011). This may be achieved by slab rollback or upper plate retreat in two dimensions, or by transtension in three dimensions.

During slab rollback the trench retreats away from the upper plate, causing it, and the accretionary wedge, to extend and thin. This creates an opportunity for buoyant high-pressure rocks to exhume into the created space (Dewey, 1980; Malusà et al., 2011). Exhumation of HP (but not necessarily UHP) rocks in locations such as the Aegean and Calabria-Appennine belts appear to be associated spatially and temporally with slab rollback events (Brun and Faccenna 2008).

Slab rollback occurrence depends on the force balance between the density of the subducting slab, the coupling strength of the subduction zone and the viscosity of the mantle lithosphere that the slab is rolling back into. Subduction velocity is also involved – slower subduction rates allow gravity to exert more of a dominating force on the slab, creating a more vertical trajectory and decoupling the slab from the upper plate (e.g. shown in numerical models such as those presented by Warren et al., 2008b). The (transient) insertion of thicker continental material into the subduction zone may trigger a decrease in the subduction rate, hence allowing the subducting slab to steepen and roll back (Brun and Faccenna, 2008; Fig. 5). The ensuing trench retreat could therefore create space for detached subducted material to exhume.

Alternatively, other numerical models suggest that extension at the surface is driven by the upwards-acting force of the exhuming plume. In cases where the exhuming material is particularly buoyant and/or moving at high velocity, it "punches" through the overlying material and forces lateral flow (Warren et al., 2008a; Beaumont et al., 2009; Butler et al., 2011; Fig. 2). In such cases subsequent re-working of the exhumed material during lateral transport may obliterate clues about its exhumation history (Butler et al., 2011).

5.2 Upper plate retreat

Extension along the subduction channel may be created by the retreat of the upper plate relative to the trench. The main difference in outcome between slab rollback vs. trench retreat is suggested to be the final structural position of the exhumed HP units (Fig. 6) – more towards the upper plate in the case of upper plate retreat and more towards the subducting plate in the case of slab rollback. Upper plate retreat has been suggested for the Eclogite Belt of the Western Alps (Malusà et al., 2011), but more work is needed to strengthen the suggestion that the final structural emplacement of the exhuming high-pressure rocks is unambiguously linkable to the relative motion between the subducting and upper plates.

Alternatively, the emplacement of the exhuming (U)HP material on either side of the accretionary wedge may be related to whether the upper plate deforms or acts as a backstop (Butler et al., 2011). Further detailed structural, geochronological and regional tectonic information is needed to determine whether either of these scenarios is likely in different subduction zone settings.

5.3 Transtension

Transtension involves both extensional and transverse shear, i.e. a combination of normal and strike-slip faulting. In this way, space is made available for exhuming material in three dimensions, whereas slab rollback and subduction zone divergence operate in two dimensions. Exhumation in a transtensive region is achieved in a similar manner to slab rollback or subduction zone divergence but diachronously along strike, as the orogen "unzips". Transtensional and rifting mechanisms have been proposed for the Miocene–Pliocene Papua New Guinea eclogites (e.g. Ellis et al., 2011) as well as the Devonian Western Gneiss region in the Norwegian Caledonides (e.g. Fossen, 2010).

In the Norwegian Caledonides, high and ultra-high-pressure rocks are exposed in a series of gentle corrugational folds beneath a major flat-lying extensional shear zone (Root et al., 2005; Fossen, 2010). Recent reconstructions suggest that Devonian-aged extension in the Caledonides was concentrated in a lozenge-shaped region, with maximum extension between the Western Gneiss Region in Norway and the Fjord Region of East Greenland (Fossen, 2010; Fig. 7a). This reconstruction suggests that the corrugational folding of the exhuming plate and focused exhumation within that region may have been caused by the transtensional tectonics (as shown by analogue experiments; Venkat-Ramani and Tikoff June, 2002). The combination of high strain and hot, partially molten rocks may have locally accelerated the exhumation of the Western Gneiss Region to at least mid-crustal levels. Later extension at higher crustal levels appears to have facilitated the final exhumation to the surface as well as creating the space for large Devonian-aged depositional basins (Fig. 7b).

5.4 Slab breakoff

The main effect of slab breakoff during on-going collision in numerical models with a fixed velocity boundary condition is to decrease the angle of subduction of the subducting plate (Boutelier et al., 2004; Warren et al., 2008c). In many numerical models of subduction and collision processes, slab breakoff is, like erosion, not required in order for exhumation to proceed. However, slab breakoff may allow a pulse of heat to reach the bottom of the slab from asthenospheric material replacing the subducting slab. This input of heat may drive melting of crustal materials, which may in turn facilitate detachment and exhumation.

5.5 Whole-scale subduction reversal: eduction

Many exhumed (U)HP terranes share characteristic structural features, including a structural dome cored by the (U)HP material, the presence of a thrust-sense structure at the base of the terrane and a normal-sense structure at the top of the terrane (e.g. Beaumont et al., 2009). Despite suggestions of the existence of a basal thrust in Norway's WGC (e.g. Andersen et al., 1991; Hacker et al., 2010), no such structure has yet been confirmed in the field. An exhumation mechanism which does not require the formation of a basal thrust (i.e. the subducted crust remains attached to its basal substrate) is that of eduction: a whole-scale reversal of subduction due to isostatic rebound following slab breakoff (Andersen et al., 1991; Duretz et al., 2012). Here the term "slab breakoff" refers to the thinning, necking and eventual separation of the subducting plate.

The term "eduction" was originally coined to explain exhumation of the Californian Franciscan blueschists during the subduction of an actively spreading ridge (Dixon and Farrar, 1980). Andersen et al. (1991) redefined the term to

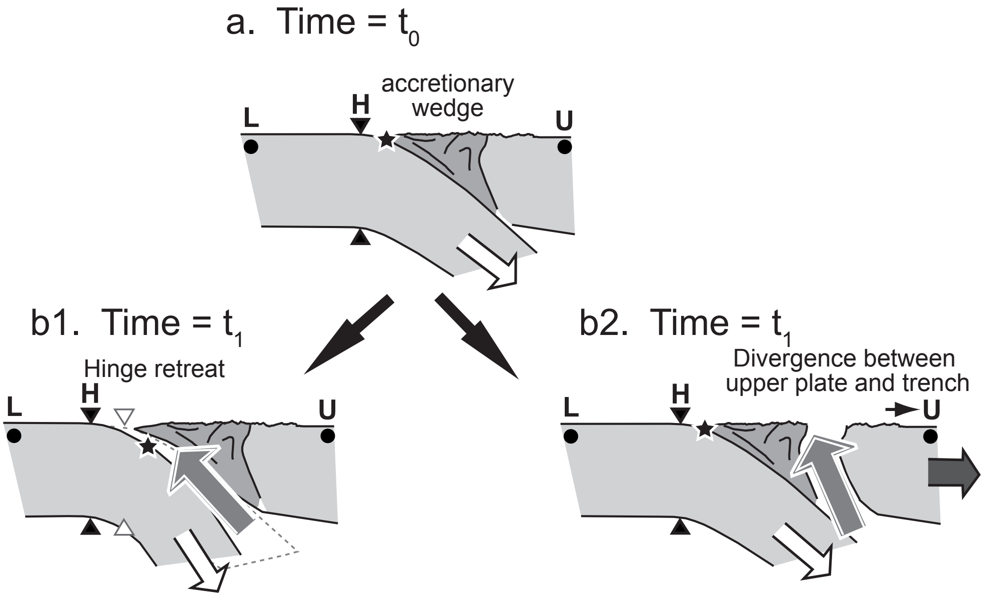

Fig. 6. Figure modified from Malusà et al. (2011) showing a suggested difference in structural position of exhumed UHP units depending on the divergence of the upper and lower plates above the exhuming high-pressure rocks. **(a)** Subduction zone before exhumation. **(b1)** Retreat of subduction hinge, with high-pressure rocks exhumed on the lower plate side of the orogen. **(b2)** Upper plate motion away from the trench, with high-pressure rocks exhumed on the upper plate side of the orogen. L = lower plate, U = upper plate; black dots indicate fixed positions on those plates. H = subduction hinge.

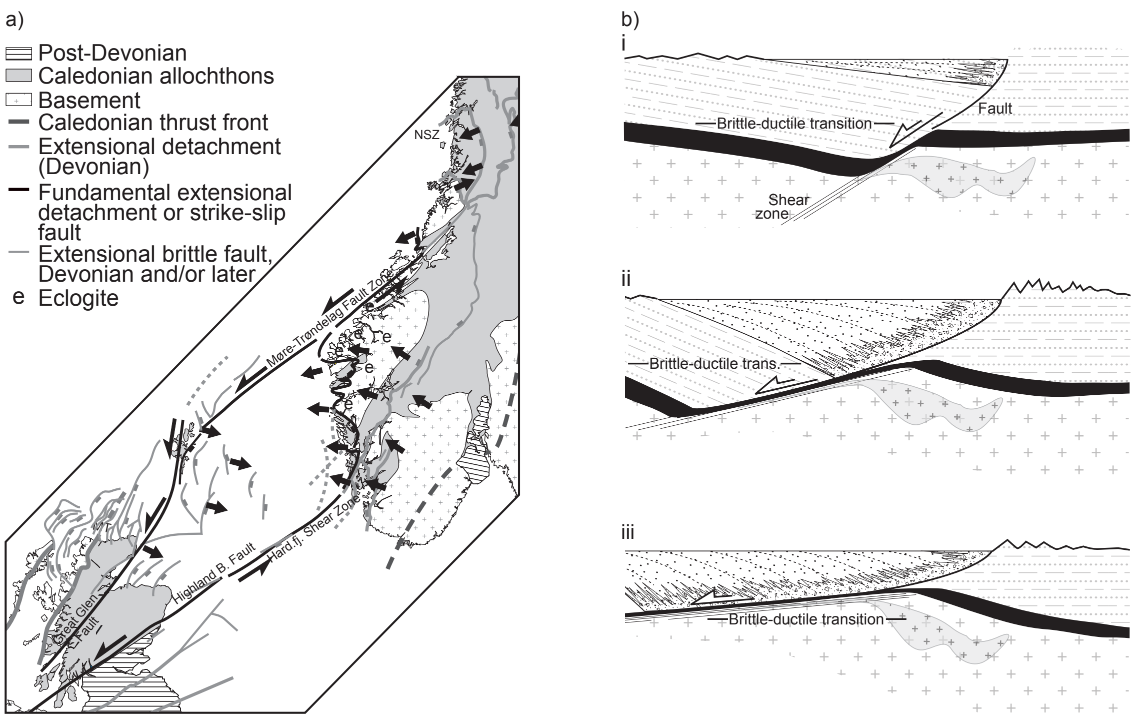

Fig. 7. Figures modified from Fossen (2010) showing the suggested role of transtension in the exhumation of the Caledonian ultra-high-pressure Western Gneiss Region in Norway. **(a)** Sketch structural map of western Norway and Scotland showing the major strike-slip and detachment faults, gross geology and location of major eclogite bodies. **(b)** Illustration of how low-angle detachments may form initially as high-angle faults that rotate to lower angles at increasing strain. This allows rotating beds to develop in hanging-wall basins, such as the Devonian Hornelen Basin.

include a period of subduction reversal following lab detachment. In this conceptual model, thinning, necking and removal of the subducting oceanic slab occurs following continental collision and partial subduction of the colliding continent. Subsequent to the removal of the dense root, the subducted, now (U)HP, continental plate exhumes coherently, driven by buoyancy, along a normal-sense structure at its top, but it remains attached to the original substrate at its base. Eduction in this sense has been demonstrated in recent numerical models where the incoming plate velocity boundary condition is turned off after a period of convergence and further convergence in the model is driven by the continuing subduction of the slab (Duretz et al., 2012). Vertical exhumation of up to 60 km is shown in the models, with initial exhumation rates on the order of $8\,\mathrm{cm\,a^{-1}}$.

6 Discussion and conclusions

Despite our understanding of crustal subduction, metamorphism and exhumation to and from depths > 100 km developing considerably since the discovery of natural coesite in crustal rocks, there are still major uncertainties about the mechanisms and processes by which crust is transported to and from mantle depths during oceanic plate subduction and continental collision. The three main types of (U)HP terranes, accretionary wedge, serpentinite subduction channel and continental, may coexist along the same subduction zone or transition from one to the other as oceanic subduction gives way to continental collision.

Clearly in many cases a combination of buoyancy and tectonic forces operate either together or separately, or at different times and places within a subduction zone, to exhume deeply subducted material. The dominant mechanism at any time or place and resulting exhumation rate in each case depends on the density difference between the exhuming material and surrounding rocks, the viscosity of the subducting and exhuming material (itself dependent on temperature, fluid concentration, melt proportion, mineralogy and microstructure), shear traction, velocity of subduction, and the coupling between the upper and subducting plates, which influences the amount of space available for surface-directed motion.

Future research needs to be directed towards determining how, when and where the force balance within a subduction or continental collision zone changes over time. Key questions that remain unresolved include the following. (1) To what extent is exhumation transitory and discrete vs. continuous and long-lived, and which tectonic environments favour the different scenarios? (2) What is the balance between buoyancy and tectonic forces in driving exhumation in different tectonic environments, and how do these change with time? (3) Which factors define the timescale(s) over which HP and UHP rocks are formed and exhumed? (4) How, where and when does exhumation change from being dominated by buoyancy, to being dominated by tectonics, to being dominated by surface processes?

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